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The lifecycle of semidiurnal internal tides over the northern Mid-Atlantic Ridge

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ABSTRACT

16 The lifecycle of semidiurnal internal tides over the Mid-Atlantic Ridge
17 (MAR) sector south of the Azores is investigated using in situ, a high-
18 resolution mooring and microstructure profiler, and satellite data, in combi-
19 nation with a theoretical model of barotropic-to-baroclinic tidal energy con-
20 version. The mooring analysis reveals that the internal-tide horizontal en-
21 ergy flux is dominated by mode 1, and that energy density is more distributed
22 among modes 1-10. Most modes are compatible with an interpretation in
23 terms of standing internal tides, suggesting that they result from interactions
24 between waves generated over the MAR. Internal tide energy is thus concen-
25 trated above the ridge and is eventually available for local diapycnal mixing,
26 as endorsed by the elevated rates of turbulent energy dissipation, ϵ , estimated
27 from microstructure measurements. A spring-neap modulation of energy den-
28 sity on the MAR is found to originate from the remote generation and radia-
29 tion of strong mode-1 internal tides from the Atlantis Meteor Seamount Com-
30 plex. Similar fortnightly variability of a factor of 2 is observed in ϵ , but this
31 signal's origin cannot be determined unambiguously. A regional tidal energy
32 budget highlights the significance of high-mode generation, with 81% of the
33 energy lost by the barotropic tide being converted into modes > 1 , and only
34 9% into mode 1. This has important implications for the fraction of local dis-
35 sipation to the total energy conversion, q , which is regionally estimated to be
36 ~ 0.5 . This result is in stark contrast with the Hawaiian Ridge system, where
37 the radiation of mode-1 internal tides accounts for 30% of the regional energy
38 conversion, and $q < 0.25$.

39 1. Introduction

40 Understanding what sets the strength and geographical variability of oceanic diapycnal mixing
41 is a critical issue in physical oceanography, because of the central role that turbulent mixing pro-
42 cesses play in the oceanic meridional overturning circulation and its impact on climate (e.g., Munk
43 and Wunsch 1998). A large fraction of the energy available for diapycnal mixing is provided by
44 the tides (Wunsch and Ferrari 2004), with satellite measurements indicating that the semidiurnal
45 M_2 barotropic tide dissipates one third of its energy in the deep ocean globally (Egbert and Ray
46 2000, 2001). This dissipation is localized to specific hotspots in which enhanced tidally-driven
47 turbulent dissipation is revealed by in situ measurements, mostly over mid-ocean ridges (Polzin
48 et al. 1997; Rudnick et al. 2003) and near isolated seamounts (Lueck and Mudge 1997). The route
49 to dissipation of the barotropic tide in the deep ocean primarily involves a conversion into baro-
50 clinic tides, i.e., internal waves with tidal frequencies. Internal tides form a reservoir of turbulent
51 energy, the dissipation of which results in irreversible diapycnal mixing. The fate of this reservoir
52 – mainly, where and how internal waves break – is poorly understood on a global scale, yet is of
53 key importance in setting the geography of diapycnal mixing (MacKinnon et al. 2017). Diapy-
54 cnal mixing is heterogeneous and strongly impacts the distributions of tracers and water masses
55 (Armi 1979) and the intensity and structure of the overturning circulation (Mashayek et al. 2015;
56 de Lavergne et al. 2016).

57 This paper addresses the lifecycle – from generation to dissipation – of semidiurnal internal tides
58 over the Mid-Atlantic Ridge (MAR) sector south of the Azores, by combining a theoretical model
59 with multi-source in situ and satellite data. Our primary goals are to document the key stages of
60 the internal tides’ lifecycle, and to outline the energy budget of the internal tides in the region.
61 The northern MAR is a relatively unexplored source of internal tides compared to the more widely

62 studied Hawaiian Ridge system [as part of the Hawaiian Ocean Mixing Experiment (HOME, e.g.,
63 Rudnick et al. 2003)] and the southern MAR [under the auspices of the Brazil Basin Tracer Release
64 Experiment (BBTRE, Polzin et al. 1997; Ledwell et al. 2000)]. However, recent theoretical (Melet
65 et al. 2013; Lefauve et al. 2015) and numerical (Timko et al. 2017) modelling studies suggest that
66 the northern MAR is an important site for internal tide generation and dissipation and, as such,
67 provides an interesting point of contrast to the Hawaiian Ridge and southern MAR.

68 The work presented here is part of the RidgeMix project, which seeks to understand and quantify
69 the upward supply of nutrients to the upper layers of the North Atlantic subtropical gyre. As part
70 of RidgeMix, a mooring was deployed on the edge of the MAR, designed to resolve variability in
71 velocity and temperature at tidal and higher frequencies throughout the entire water column with
72 high vertical resolution. This mooring provides data with which local internal tide dynamics may
73 be described for up to 10 baroclinic modes. In addition, direct measurements of turbulent energy
74 dissipation from microstructure profilers were obtained above the MAR, to assess the rate of dis-
75 sipation of internal tides. Application of a 2-D spectral model of barotropic-to-baroclinic energy
76 conversion (St. Laurent and Garrett 2002) and analyses of tidal-model estimates of barotropic tidal
77 dissipation (Egbert and Ray 2000) and satellite altimetry-derived mode-1 horizontal energy flux
78 data (Zhao et al. 2016) allow us to extend our understanding of the internal tides' lifecycle to a
79 regional scale.

80 In sections 2 and 3, we introduce the data and methods used in this study, respectively. The char-
81 acteristics of semidiurnal internal tides, characterized with a combination of a theoretical model,
82 mooring data and microstructure measurements, are presented in section 4. Our regional perspec-
83 tive of tidal energy conversion and dissipation is discussed in section 5. Our main conclusions are
84 drawn in section 6.

85 2. Data

86 In this section, we briefly describe in-situ data collection conducted during the RidgeMix cruise
87 (Sharples 2016). We then document the global gridded datasets that we use to compute tidal
88 energy-related quantities on a regional scale (section 3).

89 *a. RidgeMix data*

90 1) MOORING DATA

91 A mooring was deployed at 36.23°N, 32.75°W (Fig. 1a) on 26/09/2015 and recovered on
92 04/07/2016. It was equipped with 41 RBR self-logging thermistors, two TRDI 75-kHz Long
93 Ranger Acoustic Doppler Current Profilers (ADCPs) and two Flowquest 75-kHz ADCPs. The
94 positioning of thermistors and ADCPs along the mooring line is shown in Fig. 1b. 36 thermis-
95 tors monitored temperature with a sampling period of 15 s during the whole mooring deployment,
96 whereas 5 thermistors stopped recording between a few days and a month after deployment. The
97 spacing between thermistors was reduced where the stratification is maximum in order to capture
98 high (up to 10) dynamical modes (section 3 and Fig. 2). The ADCPs recorded hourly averaged
99 horizontal velocity (over 50 and 150 pings for the TRDI and Flowquest ADCPs, respectively) with
100 8-m vertical bins. Their positioning allowed sampling of the entire water column down to ~100 m
101 above the seafloor.

102 2) MICROSTRUCTURE DATA

103 The rate of turbulent energy dissipation ε was determined directly using vertical microstructure
104 profilers (VMPs). We deployed free-falling Rockland Scientific International ¹ (RSI) VMP-6000
105 instruments at stations on and off the ridge (see Fig. 1a for locations). A tethered RSI VMP-

¹<http://rocklandscientific.com>

2000 instrument was deployed continuously during 25-h stations (thus sampling two semidiurnal tidal cycles) in the vicinity of the mooring location during spring and neap tides (06/06/2016 and 28/06/2016, respectively). VMPs record velocity shear $\partial u/\partial z$ and temperature variance at centimeter scales. Assuming isotropy, the rate of turbulent energy dissipation is given by $\epsilon = 15\nu/2\overline{(\partial u/\partial z)^2}$ [W kg^{-1}], where ν is the molecular viscosity of seawater (Oakey 1982). In order to compare dissipation with model estimates of energy conversion, we compute the depth-integrated dissipation between 50 m and the seafloor

$$\epsilon_z = \int_{-H}^{-50 \text{ m}} \rho_0 \epsilon \, dz \, [\text{W m}^{-2}], \quad (1)$$

where H is the local depth and ρ_0 is the mean density of the profile. We did not include the uppermost 50 m, where mixed-layer processes are expected to dominate compared to internal tide breaking.

116 *b. Global gridded datasets*

117 1) SRTM30_PLUS

118 The Shuttle Radar Topography Mission dataset (SRTM30_PLUS, Becker et al. 2009) is a global
119 bathymetry dataset at a 30-sec resolution. SRTM30_PLUS is based on the 1-min Smith and
120 Sandwell (1997) bathymetry and incorporates higher resolution data from ship soundings where
121 ever available. The MAR sector south of the Azores has been intensively surveyed (see Fig. 3
122 in Timko et al. 2017), and SRTM30_PLUS is significantly enriched by small-scale topographic
123 features compared to the Smith and Sandwell (1997) dataset.

124 2) WOA13

125 Temperature and salinity data required to compute the buoyancy frequency are from the 1°-
126 resolution World Ocean Atlas 2013 version 2 climatology² (WOA13, Locarnini et al. 2013; Zweng
127 et al. 2013). This climatology is computed by objective analysis of historical hydrographic profiles
128 from many different sources.

129 3) TPXO

130 Barotropic-tide currents (amplitude and phase) were extracted from the 1/12°-resolution inverse
131 tidal model for the Atlantic Ocean, the TPXO AO_ATLAS,³ a regional version of TPXO8 (Egbert
132 and Erofeeva 2002). We hereafter refer to this dataset as TPXO.

133 4) MODE-1 M₂ ENERGY FLUXES AND SEA SURFACE HEIGHT FROM SATELLITE ALTIMETRY

134 Mode-1 M₂ internal-tide horizontal energy flux and sea-surface height (SSH) data at a horizontal
135 resolution of 1/5° from Zhao et al. (2016) were used in this study to quantify the propagation of
136 baroclinic tidal energy on a regional scale. Zhao et al. (2016) use a two-dimensional plane wave fit
137 method to extract internal tides from satellite SSH and apply a modal decomposition that allows the
138 inference of mode-1 internal tide pressure from SSH. Assuming that the energy partition between
139 potential and kinetic energy components depends only on latitude and tidal frequency, the internal
140 tide velocity is also estimated from SSH. Finally, vertically-integrated horizontal energy fluxes,
141 \mathbf{F}_s^1 , are computed (Appendix A in Zhao et al. 2016). Positive divergence of the horizontal energy
142 flux, hereafter denoted $(\nabla \cdot \mathbf{F}_s^1)^+$, indicates regions of mode-1 internal tide generation.

²<https://www.nodc.noaa.gov/OC5/woa13/>

³<http://volkov.oce.orst.edu/tides/AO.html>

3. Methods

In this section, we outline the methodology used in this study. First, we implement a theoretical model for the generation of internal tides (section 3a). This is followed with an analysis of mooring data to characterize internal tide properties (section 3b). Finally, we estimate the barotropic tidal energy loss on a regional scale from a tidal model (section 3c).

a. Theoretical model of barotropic-to-baroclinic energy conversion

In a stratified fluid, the interaction of a current with varying topography generates internal waves. Under different sets of flow characteristics and dynamical assumptions, models and parameterizations for internal wave generation have been developed (e.g., Baines 1998; Nycander 2005). When the current of interest is the barotropic tide, two dimensionless parameters mainly govern the nature of internal waves (St. Laurent and Garrett 2002; Garrett and Kunze 2007): the ratio of topographic slope, $s = \nabla h$, to wave characteristic slope, $\alpha = \sqrt{(\omega^2 - f^2)/(N^2 - \omega^2)}$; and the ratio of tidal excursion to topographic length scale, ku_0/ω , where u_0 is the barotropic tidal velocity and k is the topographic horizontal wavenumber. In the deep ocean – i.e., far from continental shelves and slopes –, the major topographic features responsible for barotropic tidal dissipation are mid-ocean ridges (Egbert and Ray 2000, 2001). Over these ridges, topographic slopes are dominantly subcritical ($s/\alpha < 1$) and tidal excursions are smaller than topographic scales ($ku_0/\omega < 1$), due to weak barotropic tidal currents [$u_0 = O(1) \text{ cm s}^{-1}$]. Most deep-ocean barotropic-to-baroclinic energy conversion models are based on these two assumptions, which permit the derivation of linear equations (e.g., Bell 1975a,b; Jayne and St Laurent 2001; Llewellyn Smith and Young 2002; St. Laurent and Garrett 2002; Nycander 2005).

Among the various existing models, we chose to use a two-dimensional spectral model that follows St. Laurent and Garrett (2002). Although computationally more expensive, this method

166 offers an extensive characterization of the vertical energy flux, providing information such as
 167 modal content and flux direction. The barotropic-to-baroclinic vertical energy flux E_f [see Eq. (10)
 168 in St. Laurent and Garrett (2002)], may be estimated as

$$E_f(K, \theta) = \frac{1}{2} \rho_0 \frac{[(N_b^2 - \omega^2)(\omega^2 - f^2)]^{1/2}}{\omega} \times (u_e^2 \cos^2 \theta + v_e^2 \sin^2 \theta) K \phi(K, \theta) \quad (2)$$

$$[\text{W m}^{-2}(\text{rad m}^{-1})^{-2}],$$

169 where N_b is the buoyancy frequency close to the bottom computed from WOA13; u_e (v_e) is
 170 the barotropic tidal velocity amplitude from TPXO, in the direction of the semimajor (semimi-
 171 nor) axis of the tidal ellipse $[(x_e, y_e)$ coordinate system]; $K = (k_x^2 + k_y^2)^{1/2}$ is the total horizon-
 172 tal wavenumber, with k_x and k_y being the horizontal wavenumbers in the (x_e, y_e) coordinate sys-
 173 tem, and $\theta = \arctan(k_y/k_x)$. The 2-D power spectrum of topography, ϕ , is normalized to satisfy
 174 $\int_0^{2\pi} \int_0^\infty \phi(K, \theta) K dK d\theta = \overline{h^2}$, where $\overline{h^2}$ is the mean square height of topography.

175 From Eq. (2), we define the azimuthally-averaged vertical energy flux as

$$E_f^a(K) = \frac{1}{2\pi} \int_0^{2\pi} E_f(K, \theta) K d\theta [\text{W m}^{-2}(\text{rad m}^{-1})^{-1}], \quad (3)$$

176 and the radially-integrated vertical energy flux as

$$E_f^r(\theta) = \int_{K_1}^\infty E_f(K, \theta) dK [\text{W m}^{-2}(\text{rad m}^{-1})^{-1}]. \quad (4)$$

177 The total vertical energy flux is

$$E_f^t = \int_0^{2\pi} \int_{K_1}^\infty E_f(K, \theta) K dK d\theta [\text{W m}^{-2}], \quad (5)$$

178 where the lower boundary of integration in wavenumber space is the mode-1 equivalent wavenum-
 179 ber, K_1 , to take into account the finite depth of the ocean (Llewellyn Smith and Young 2002). We
 180 also define the vertical energy flux into mode j as

$$E_f^j = \int_0^{2\pi} \int_{K_j - \delta K/2}^{K_j + \delta K/2} E_f(K, \theta) K dK d\theta [\text{W m}^{-2}], \quad (6)$$

181 where $\delta K = K_2 - K_1$, and the equivalent wavenumber of mode j is

$$K_j = \frac{j\pi(\omega^2 - f^2)^{1/2}}{N_0 b}. \quad (7)$$

182 N_0 and b are parameters of an exponential fit to the buoyancy frequency $N = N_0 \exp(z/b)$ (St. Lau-
183 rent and Garrett 2002).

184 *b. Energy density and horizontal energy flux from mooring data*

185 Internal-tide energy density, E , and horizontal energy flux, \mathbf{F} , are estimated from mooring data
186 following Nash et al. (2005). Here, we briefly recall the main steps of their procedure.

187 The wave velocity, $\mathbf{u}'(z, t)$, is defined as

$$\mathbf{u}'(z, t) = \mathbf{u}(z, t) - \bar{\mathbf{u}}(z) - \bar{\mathbf{u}}_0(t), \quad (8)$$

188 where $\mathbf{u}(z, t)$ is the instantaneous velocity as recorded by the instrument, $\bar{\mathbf{u}}(z)$ is the time mean of
189 that velocity, and $\bar{\mathbf{u}}_0(t)$ is defined by the baroclinicity condition $\frac{1}{H} \int_{-H}^0 \mathbf{u}'(z, t) dz = 0$. Here, the
190 time-mean velocity is defined as the 5-day running mean (as in Zhao et al. 2010) to filter out meso-
191 to submesoscale processes (at least below the surface mixed layer). Sensitivity on the length of the
192 time window has been tested and found to be weak as the signals are further band-passed filtered
193 in the semidiurnal waveband.

194 The wave pressure, $p'(z, t)$, is defined as

$$p'(z, t) = p_{\text{surf}}(t) + \int_z^0 \rho'(\hat{z}, t) g d\hat{z}, \quad (9)$$

195 where p_{surf} is the surface pressure, g is the acceleration of gravity, and ρ' is the density pertur-
196 bation associated with the wave. Although p_{surf} is not measured, $p'(z, t)$ is constrained by the
197 baroclinicity condition, $\frac{1}{H} \int_{-H}^0 p'(z, t) dz = 0$. Formally, the density perturbation ρ' is defined as

$$\rho'(z, t) = \rho(z, t) - \bar{\rho}(z), \quad (10)$$

198 where $\rho(z, t)$ is the instantaneously measured density and $\bar{\rho}(z)$ is the time-mean vertical density
 199 profile. Since the mooring is only equipped with thermistors, we cannot derive the density di-
 200 rectly. As an alternative, $\rho'(z, t)$ is inferred from the vertical displacement of isopycnals $\xi(z, t)$
 201 (approximated as isotherms here) relative to their mean position:

$$\rho'(z, t) = (\bar{\rho}(z)/g)\bar{N}^2(z)\xi(z, t), \quad (11)$$

202 where $\bar{N}^2(z) = -(g/\rho_0)\partial_z\bar{\rho}$ is the time-mean buoyancy frequency profile computed from WOA13
 203 temperature and salinity interpolated to the mooring position. The linear relationship between ρ'
 204 and ξ is valid due to the slowly varying profile of time-mean density with depth (Desaubies and
 205 Gregg 1981). The vertical displacement of isopycnals, $\xi(z, t)$, is given by

$$\xi(z, t) = T'(z, t)[\partial_z\bar{T}(z)]^{-1}, \quad (12)$$

206 where $T'(z, t) = T(z, t) - \bar{T}(z)$ is the temperature anomaly relative to a 5-day running mean, and
 207 $\partial_z\bar{T}(z)$ is the time-mean vertical gradient of temperature (e.g., Alford 2003). We checked that
 208 $\partial_z\bar{T}(z)$ close to the bottom was bounded by a lower value ($9 \times 10^{-4} \text{ }^\circ\text{C m}^{-1}$) representative of a
 209 stratified environment, and that would not lead to singularities in Eq. (12).

210 The variables \mathbf{u}' , p' and ξ are then filtered at the M_2 frequency, ω , using a band-pass fourth-
 211 order Butterworth filter in the bandwidth $\{c^{-1}\omega, c\omega\}$ with $c = 1.25$ (Alford 2003; Alford and
 212 Zhao 2007a; Zhao et al. 2010). We ensured that at the mooring latitude the value of c does not
 213 lead to overlapping of the waveband with the near-inertial band $\{c^{-1}f, cf\}$, where f is the Coriolis
 214 frequency. However, semidiurnal frequencies M_2 and S_2 are too close to be adequately resolved
 215 by a band-pass filtering method. The filtered variables thus contain both M_2 and S_2 , and hence
 216 display spring-neap variability.

Variables are next projected onto baroclinic modes. The baroclinic modes for vertical displacement $\Phi_n(z)$ ($n > 0$) are defined as the solutions of the eigenvalue problem

$$\frac{d^2 \Phi_n}{dz^2} + \frac{N^2(z)}{c_n^2} \Phi_n(z) = 0, \quad (13)$$

with boundary conditions $\Phi_n(0) = \Phi_n(-H) = 0$, where n is the mode number and c_n is its eigen-speed (Gill 1982), defined as

$$c_n = \frac{H}{n\pi} \int_{-H}^0 N(z) dz. \quad (14)$$

The corresponding modes for pressure and horizontal velocity, $\Pi_n(z)$, are defined as

$$\Pi_n(z) = \rho_0 c_n^2 \frac{d\Phi_n}{dz}. \quad (15)$$

The buoyancy frequency $N(z)$, computed from WOA13, and the corresponding modes $\Pi_n(z)$ for $n = 1, \dots, 10$ are shown in Fig. 2. The modes as observed by the array of thermistors are superimposed in red. Projection of variables onto these modes – e.g., for velocity, $\mathbf{u}'(z, t) = \sum_{n=0}^{10} \mathbf{u}'_n(t) \Pi_n(z)$ – uses a least-square fit method (Alford 2003; Nash et al. 2005; Zhao et al. 2010).

Combining \mathbf{u}' , p' and ξ (indices referring to modes are omitted in the following) allows computation of the depth-integrated baroclinic kinetic (KE) and potential (PE) energy densities:

$$\text{KE} = \left\langle \frac{1}{2} \int_{-H}^0 \rho (u'^2 + v'^2) dz \right\rangle [\text{J m}^{-2}], \quad (16)$$

$$\text{PE} = \left\langle \frac{1}{2} \int_{-H}^0 \rho (N^2 \xi^2) dz \right\rangle [\text{J m}^{-2}], \quad (17)$$

as well as the horizontal energy flux \mathbf{F} :

$$\mathbf{F} = \left\langle \int_{-H}^0 \mathbf{u}' p' dz \right\rangle [\text{W m}^{-1}]. \quad (18)$$

In Eqs. (16)-(18), $\langle \cdot \rangle$ denotes an average over a wave period (M_2 here).

231 *c. Barotropic tide energy loss using a tidal model*

232 Following Egbert and Ray (2000), we compute the dissipation rate of the barotropic tide D as

$$D = W - \nabla \cdot \mathbf{P} \text{ [W m}^{-2}\text{]}, \quad (19)$$

233 where W is the work done by the barotropic tide and \mathbf{P} is the barotropic tide energy flux. \mathbf{P} is
234 defined as

$$\mathbf{P} = \rho_0 g \langle \mathbf{U} \zeta \rangle, \quad (20)$$

235 where ζ is the tidal elevation and \mathbf{U} is the barotropic tide volume transport, both extracted from
236 TPXO. W is defined as

$$W = \rho_0 g \langle \mathbf{U} \cdot \nabla (\zeta_{\text{eq}} + \zeta_{\text{sal}}) \rangle, \quad (21)$$

237 where ζ_{eq} is the equilibrium tidal elevation and ζ_{sal} is the tidal elevation induced by the tide's
238 self-attraction and loading (Ray 1998).

239 **4. Structure of the semidiurnal internal tide**

240 In this section, we use a range of measurements and a theoretical model to assess the lifecycle
241 of internal tides over the MAR at the location of the RidgeMix mooring – from generation (sec-
242 tion 4a) to propagation (section 4b) and dissipation (section 4c) – before offering a summary of
243 this local perspective (section 4d).

244 *a. Theoretical estimates of internal tide generation*

245 Figure 3 illustrates the method used to estimate barotropic-to-baroclinic energy conversion (sec-
246 tion 3a) at the specific mooring location. First, the method interpolates the barotropic tidal el-
247 lipse from TPXO at the point of interest and extracts topography from SRTM30 PLUS around it
248 (Fig. 3a). Second, topography is rotated along the ellipse's axes (Fig. 3b) and its two-dimensional

power spectrum ϕ is computed (Fig. 3c). Third, ϕ is directionally weighted by tidal currents and multiplied by a factor depending on the three frequencies of the system (f , ω and N_b) to give the vertical energy flux $E_f(K, \theta)$ [Eq. (2) and Fig. 3d]. Finally, E_f is azimuthally-averaged to get its distribution as a function of horizontal wavenumber (Fig. 3e) – or equivalently its modal distribution [Eq. (7)]. Its cumulative sum eventually gives the total energy conversion (Fig. 3f). Alternatively, E_f can be integrated in the wavenumber direction to get its azimuthal dependence (Fig. 3g).

The model predicts an energy conversion at the mooring site that spans a wide range of equivalent horizontal wavenumbers, noticeably exhibiting a plateau between modes 1 and 5 and then gradually decreasing (Fig. 3e). Indeed, the rough topography of the MAR varies strongly on a wide range of scales, down to abyssal hill scales of $O(1)$ km (Goff 1991). As a consequence, high-mode internal tides are expected to be radiated, as observed (St Laurent and Nash 2004) and modeled (Zilberman et al. 2009) on the flanks of the MAR in the Brazil Basin. Superimposed on the theoretical model estimates is the spectrum of mooring-derived horizontal energy flux, converted to a vertical flux by multiplying by α , the wave characteristic slope, and dividing by the water depth. The energy flux is cut at mode 35, which is in theory the highest mode that can be resolved with 36 independent thermistors.⁴ It shows a good agreement with the theoretical model for modes higher than 5, but overestimates energy fluxes in modes 1-4 (Fig. 3e). Nonetheless, we do not expect a perfect match, as the mooring detects fluxes from remote sources – most likely propagating low modes – that are not taken into account in the model. The total vertical energy flux is 4.5 mW m^{-2} in the model, and 9.1 mW m^{-2} in the mooring data. A factor-of-two discrepancy is also found by St Laurent and Nash (2004). This may also relate to the model's failure to take

⁴This assumes that the spacing between thermistors is perfectly designed to capture the vertical structure of high modes (Fig. 2), which may not be the case for the highest modes.

271 account of the sub-tidal circulation, which could introduce variability in internal tide generation
272 (Kerry et al. 2014).

273 The model predicts the direction of the flux modulo 180° (Fig. 3g). The two preferential direc-
274 tions are almost perpendicular to the tidal ellipse's major axis ($\theta = 0$) and coincides roughly with
275 the cross-fracture zone direction (Fig. 3a). The mooring-derived flux is mainly to the southeast
276 and roughly fits in the prediction of the model (Fig. 3g). Again, we do not expect a perfect match
277 since the model is local and can not take into account remote modulation of the flux.

278 *b. Internal tide properties from high-resolution mooring data*

279 1) ENERGY DENSITY AND ENERGY FLUX

280 Time series of energy density, $E = KE + PE$, and horizontal energy flux, $F = ||\mathbf{F}||$, in modes
281 1-10 are shown in Figs. 4d,e. The energy density is mostly contained in mode 1, and gradually
282 decreases with increasing mode number (Fig. 4d). The horizontal energy flux is, on the other hand,
283 overwhelmed by mode 1, which is almost indistinguishable from the total energy flux (Fig. 4e).
284 This picture is consistent with open-ocean mooring estimates of energy density and energy fluxes
285 from the Internal Waves Across the Pacific experiment (IWAP, Zhao et al. 2010). Indeed, on
286 the one hand, the wave velocity \mathbf{u}' and displacement ξ' project qualitatively onto a few modes,
287 between 1 and 10 (not shown). On the other hand, the wave pressure p' results from the vertical
288 integration of ξ' , and is hence smoother, thus is dominated by low modes. As a consequence, the
289 kinetic [Eq. (16), Fig. 4b] and potential energy [Eq. (17), Fig. 4c] computed from \mathbf{u}' and ξ' have
290 some contributions from modes 1-10. In contrast, the horizontal energy flux [Eq. (18), Fig. 4e]
291 computed from \mathbf{u}' and p' is strongly dominated by mode 1. The time-mean and standard deviation
292 of E and F as a function of mode confirm this distribution (Figs. 5a,b and Table 1). The mode-1
293 energy flux accounts for 83% of the energy flux of modes 1-10. However, mode-1 energy density

accounts for only 45% of the energy density of modes 1-10. Our basic interpretation is that, although the bulk of – potentially – propagating energy is in mode 1, mode 2 and above (modes higher than 10 are partially captured by the mooring) contain at least 55% of the energy ultimately available for local mixing.

The robustness and steadiness of mode-1 flux compared to higher modes is also demonstrated by the time series of their direction (Fig. 4f). The mode-1 flux is always directed between east and south directions and varies slowly, likely influenced by the surrounding mesoscale eddy field (Rainville and Pinkel 2006; Dunphy et al. 2017). On the other hand, the mode-2 and -3 flux directions vary through all azimuths on daily time scales. A similar variability is found for modes greater than 3 (not shown). This short time scale variability might be attributed to interferences between waves arising from different sources, reflection and scattering (e.g., Zaron and Egbert 2014). Two-dimensional histograms of modal horizontal energy fluxes further confirm the multi-directional nature of fluxes for modes greater than 1 (Fig. 6). This high directional variability is probably linked to the multiple sources of internal tides on the MAR around the mooring. The recent comparison of mode-1 and -2 horizontal energy fluxes from a high-resolution numerical model and historical moorings further demonstrates a poorer correlation and a higher variability in mode-2 fluxes compared to mode-1 fluxes (Ansong et al. 2017).

2) GROUP VELOCITY

Following the method of Alford et al. (2006) and Alford and Zhao (2007b), we compute the group velocity of each mode from mooring estimates of energy density and horizontal energy flux $c_g^m = F/E$. The method exploits the strong correlation between E and F (scatter plots in Figs. 7a,c). Briefly, the mean energy and standard deviation are first estimated in each energy-flux bin (we chose 10 evenly-spaced bins between extreme flux values). The slope, i.e. c_g^m , and its 95%

confidence interval are then computed by linear regression. Probability density functions (PDFs) of the populations of F/E also give an overview of the distribution (Figs. 7b,d). Estimates of c_g^m are compared to theoretical values of group velocity for freely propagating waves:

$$c_g = c_n \frac{(\omega^2 - f^2)^{1/2}}{\omega}, \quad (22)$$

where c_n is the mode- n eigenspeed [Eq. (14)]. Alford and Zhao (2007b) also developed a simple model for the perceived group velocity of a standing wave resulting from the interaction of two waves propagating in the opposite direction (see also Nash et al. 2004). This perceived group speed c_g^s is a spatial modulation of c_g ,

$$c_g^s = \frac{2\omega f \sin(2kx)}{\omega^2 - f^2 \cos(2kx)} c_g, \quad (23)$$

where k is the wavenumber in the x direction (see Appendix in Alford and Zhao 2007b). In the following, c_g^s refers to the mean group velocity over one wavelength.

The estimated mode-1 group velocity ($1.09 \pm 0.10 \text{ m s}^{-1}$) agrees particularly well with the group velocity of a standing wave (1.08 m s^{-1} , Fig. 7a and Table 1). Interestingly, the peaks of the bimodal-like shape of the PDF of F/E coincide with c_g and c_g^s (Fig. 7b). This suggests that, although the mode-1 wave is most of the time consistent with a standing wave, specific events are more compatible with a propagating wave.

The mode-2 group velocity shows a different picture, being inconsistent with both propagating and standing wave velocities (Fig. 7c). Estimates of c_g^m are smaller than c_g and c_g^s by 48% and 37%, respectively. Such discrepancies in mode-2 group velocities with theoretical estimates are also reported over the MAR in Alford and Zhao (2007b). They attribute this slow apparent propagation to the multidirectional fluxes – observed for modes greater than or equal to 2 here (Figs. 6b-f) – that decohere the waves.

We applied the same technique to modes 1-10 and report the estimated group velocity with their 95% confidence interval in Fig. 8. Apart from mode 4, which is more consistent with a propagating wave, all modes are either more compatible with standing waves or have even smaller group velocities than expected from a standing wave. Modes greater than 8 have very small group velocity due to vanishing fluxes, and their velocities thus gradually depart from theoretical values.

3) SPRING-NEAP CYCLE

The energy density and horizontal energy flux both display a remarkable spring-neap cycle, mostly dominated by mode 1 (Figs. 4d,e). This spring-neap cycle is obviously related to the astronomical forcing, as seen in barotropic kinetic energy $KE_{bt} = \int_{-H}^0 \frac{1}{2} \rho ||\mathbf{u}||^2 dz$ (Fig. 4a). Time series of KE_{bt} from the mooring shows a good agreement with a synthetic estimate from the combination of M_2 and S_2 computed from TPXO (red line in Fig. 4a). Major peaks at the end of September and October might be associated with other long-term astronomical forcing frequencies that amplify the semidiurnal signal.

The time lag between KE_{bt} and E is estimated in lag-correlating time series, prior band-passed filtered at the spring-neap cycle (Alford and Zhao 2007a). The maximum correlation is 0.69 and is reached for a 3.4-day lag. We conjecture that the spring-neap variability – mostly seen in mode 1 – is triggered by remotely generated mode-1 internal tides that propagate up to the mooring site. In order to track down the origin of these waves, we make use of the Zhao et al. (2016) data set, which decomposes the internal tide properties (SSH and horizontal energy fluxes) into their northbound and southbound components (Figs. 9a,b). Notice that this data set contains only M_2 internal tides whereas the mooring analysis contains all semidiurnal constituents. However, we checked that the M_2 surface-tide kinetic energy dominates over other semidiurnal constituents by an order of magnitude regionally (not shown), so we expect M_2 to also dominate the internal-wave field. The

Zhao et al. (2016) data set reveals that the Atlantis Meteor Seamount Complex [green contours in
 Figs. 9a,b; see also Fig. 1 in Searle (1987) for a wider geographical setting] is a regional hotspot
 for mode-1 internal tide generation. In particular, a northbound beam emanates from the Hyères
 Seamount (31.3°N,28.9°W; green star in Fig. 9a) and points toward the mooring site, following
 the orange line in Fig. 9a. SSH interpolated along this line shows a clear oscillating signal with a
 decreasing amplitude along the path (Fig. 9c). The travelling time $t(x)$, as a function of distance
 from the source x , for this semidiurnal mode-1 internal tide is estimated as

$$t(x) = \int_{x_s}^x \frac{x'}{c_g(x')} dx', \quad (24)$$

where x_s is the seamount coordinate and c_g is the mode-1 group velocity defined in Eq. (22).
 Figure 9d shows the spatial variability of c_g – mostly depending on the bathymetry (Fig. 9c) – and
 the travelling time throughout the propagation. In theory, the internal tide reaches the mooring
 site in ~ 4 days, which is comparable to the 3.4-day lag between the astronomical forcing and the
 oceanic response. As such, the internal tide generated at the Hyères Seamount is a good candidate
 to explain the spring-neap modulation of energy density and horizontal energy flux measured at the
 mooring site. Notice that its energy flux is roughly in the opposite direction to the flux diagnosed
 at the mooring site. Hence, the superposition of the two waves is coherent with the diagnosed
 standing group velocity at the mooring site.

c. Local dissipation from microstructure measurements

Two 25-h stations with continuous tethered-VMP deployments were carried out in the vicinity of
 the mooring site during spring and neap tides (section 2a). Mean profiles of the turbulent dissipa-
 tion rate, ε , and the PDF of $\log(\varepsilon)$ for both series of casts are shown in Fig. 10. There is evidence
 for intensified dissipation during spring tide, as highlighted by the spring-tide PDF of $\log(\varepsilon)$ be-

ing skewed towards higher values compared to the neap-tide PDF (Fig. 10b). Vertical profiles of ϵ also reveal a higher spring-time dissipation at almost all depths with enhanced differences in the bottom-most 500 m (Fig. 10a). In this depth range, ϵ reaches $10^{-9} \text{ W kg}^{-1}$, as routinely observed over rough topography of the world's oceans (Kunze 2017). Notice that the tethered VMP could not dive deeper than ~ 400 m above the seafloor (~ 2200 m) due to wire length limitations, and we expect dissipation to further increase with depth in excess of 1800 m.

The depth-integrated dissipation ϵ_z [Eq. (1)] is $1.3 \pm 1.1 \text{ mW m}^{-2}$ during spring tide, and $0.7 \pm 0.4 \text{ mW m}^{-2}$ during neap tide. A similar factor-of-two difference between spring- and neap-tide dissipation has been observed on the Hawaiian Ridge (Klymak et al. 2006). Notice that ϵ_z is likely to be underestimated due to undersampling of the water column. Nonetheless, these high levels of dissipation may be due to the enhanced local generation of high-mode internal tides that carry most of the shear variance (Fig. 5c) and are prone to rapid breaking close to their generation site (in a similar fashion as on the Oregon continental slope, Nash et al. 2007). In addition, the spring-neap modulation and bottom intensification of dissipation suggests that the elevated turbulence may be triggered by a direct breaking of high-mode internal tides (Klymak et al. 2008). Note, however, that we are unable to verify that the spring-neap component of dissipation is phase-locked with astronomical forcing.

d. Summary of the local perspective

In summary, the high-resolution mooring data provide us with a detailed insight into internal tide dynamics on the northern MAR. The horizontal energy flux is highly dominated by mode 1 (0.83 kW m^{-1}), and is rather steady in direction. Its intensity displays a strong spring-neap cycle lagging by 3.4 days from the astronomical forcing, hence pointing to a modulation by remote sources. The Hyères Seamount – a hotspot of mode-1 internal tide generation of the Atlantis

404 Meteor Seamount Complex – is a very likely candidate as it radiates an internal-tide beam towards
405 the mooring site, whose travelling time is close to the spring-neap cycle lag to the astronomical
406 forcing.

407 The horizontal energy fluxes associated with modes 2-10 are very weak ($<0.07 \text{ kW m}^{-1}$) and
408 vary strongly in direction, likely due to the interactions of waves generated by numerous, dis-
409 tributed sources on the MAR. In turn, the energy density is more widely partitioned between
410 modes, with mode 1 accounting for a smaller fraction of the total energy density than the sum of
411 modes 2-10 (0.84 vs 0.95 kJ m^{-2}). Examination of the propagation velocity revealed that most
412 of the modes are compatible with standing waves. This implies that internal-tide energy is likely
413 to remain concentrated over the MAR, and thereby become ultimately available for near-local tur-
414 bulent mixing. In line with this result, microstructure measurements performed at the mooring
415 site reveal elevated and bottom-intensified turbulent energy dissipation. The energy conversion
416 model further confirms that high modes are expected to be generated. The model possibly under-
417 estimates conversion into low modes – although low modes diagnosed from mooring data may
418 originate from remote sources (Figs. 9a,b) – but its agreement with mooring-derived fluxes for
419 modes greater than 3 is remarkable.

420 **5. Regional perspective**

421 In order to get a broader view of internal tide dynamics over the northern MAR, we performed
422 a regional energy budget using different data sources. The barotropic tide energy loss and internal
423 dissipation should be equal in the absence of energy transport by internal tides. However, low-
424 mode internal tides play a role in redistributing energy. In addition, energy entering low modes
425 does not dissipate locally. In the following, the barotropic tide energy loss, D , is estimated via
426 a tidal model (section 3c); the tidal barotropic-to-baroclinic conversion, E_f , is estimated via a

two-dimensional spectral model (section 3a); and the conversion to mode-1 internal tide is also estimated via satellite altimetry $(\nabla \cdot \mathbf{F}_s^1)^+$ (section 2b and Zhao et al. 2016). For this exercise, we extended the conversion model to a regional domain spanning from 22°N (southern edge of the RidgeMix cruise) to 42°N, slightly north of the Azores. Using global data sets described in section 2b, we computed E_f on a regular 1/4° grid.

The regional distribution of the total energy conversion, E_f^t , the mode-1 energy conversion, E_f^1 , the energy conversion into modes ≥ 2 , $E_f^{2-\infty}$, the barotropic tide energy loss, D and the satellite-estimate of mode-1 energy conversion, $(\nabla \cdot \mathbf{F}_s^1)^+$, are shown in Figs. 11a-e. The highest levels of conversion ($>10 \text{ mW m}^{-2}$) are mostly found at depths shallower than 2000 m near the Azores and the Atlantis Meteor Seamount Complex. This is due to strong barotropic currents and increased bottom stratification associated with shallower depths. The regions of strong barotropic tide energy loss are collocated with these areas, although they are more spatially widespread around the Atlantis Meteor Seamount Complex. Other hotspots of conversion ($>5 \text{ mW m}^{-2}$) are found on the edge of the MAR. This is where topography is roughest, thus contributing to a rich energy conversion through a broad range of scales as highlighted in section 4.

The energy conversion into mode 1 agrees well between the two independent estimates (Figs. 11b,e). As shown above, the Atlantis Meteor Seamount Complex is the main source of mode-1 internal tides (see Figs. 9a,b). Another hotspot is the Azores Islands, which the satellite product misses likely due to the proximity of land. Importantly, both products concur on a very low mode-1 conversion at the MAR ($<1 \text{ mW m}^{-2}$). In contrast, strong generation of modes ≥ 2 occurs on the MAR, and accounts for most of the energy conversion (compare Figs. 11a and 11c). Using a different method for estimating energy conversion into normal modes, Falahat et al. (2014b) demonstrate a qualitatively similar distribution (their Fig. 6).

450 All quantities are further summed over the hatch-free area in Figs. 11a-e, where the ocean depth
451 lies between 200 and 4000 m. This area isolates the MAR and does not include the Azores plateau,
452 where the assumption of small tidal excursion is likely to be violated. Two robust conclusions can
453 be drawn from this budget (Fig. 11f). First, the close agreement between D (16.4 GW) and E_f^t
454 (13.7 GW) confirms previous assumptions that most of the energy dissipated by the barotropic tide
455 in the deep ocean is converted into internal tides – 84% here – and not dissipated by bottom friction
456 like on continental shelves (e.g., Egbert and Ray 2000). Second, more specific to the northern
457 MAR, energy conversion into mode 1 only accounts for 9% (1.2 GW) of the total conversion
458 (7% of the barotropic tide energy loss) and higher modes thus represent the bulk of the energy
459 conversion (12.4 GW, 81%). The satellite product confirms the modest contribution of mode 1
460 (1.7 GW).

461 As a point of comparison, the Hawaiian Ridge system dissipates 20 GW of barotropic tidal
462 energy (Egbert and Ray 2001), of which 6 GW (30%) is converted into mode 1 (Merrifield and
463 Holloway 2002). The difference between the distribution of energy stems from the different to-
464 pographic shapes of the two ridge systems. The Hawaiian Ridge has abrupt flanks that generate
465 intense mode-1 tides, which may propagate far away from the ridge (Zhao et al. 2010). In contrast,
466 the MAR has a wider rift valley (in the fracture zone direction, the direction perpendicular to the
467 ridge edge) and hosts taller and steeper abyssal hills due to its slow spreading rate (Goff 1991).
468 The latter are known to generate high-mode internal tides (Melet et al. 2013; Lefauve et al. 2015;
469 Timko et al. 2017) that are prone to rapid breaking.

470 The VMP data allow us to gain some insight into the distribution in turbulent dissipation levels
471 across and beyond the MAR (Fig. 12). The most notable feature is the strong on- vs. off-ridge
472 contrast, with increased dissipation occurring above the rough topography of the MAR [as also
473 evidenced in the Brazil Basin by Polzin et al. (1997) and Ledwell et al. (2000)]. Point-wise dissi-

474 pation rate is often $\geq 10^{-9} \text{ W kg}^{-1}$ over the ridge, and decays to $O(10^{-11} - 10^{-10}) \text{ W kg}^{-1}$ off
 475 the ridge (Fig. 12b). The vertical distribution of ε is beyond the scope of this study, and we focus
 476 on the depth-integrated dissipation ε_z from 50 m (to exclude turbulence related to mixed-layer
 477 processes) to the seafloor [Eq. (1) and Fig. 12a]. ε_z and the local energy conversion, E_f^t , exhibit
 478 similar patterns, attaining maximum values on the ridge and minimum rates off the ridge. Note
 479 that ε_z is smaller than E_f^t everywhere, which is expected since a fraction of energy may radiate
 480 away.

481 The fraction of the local dissipation to the total energy conversion, $q = \varepsilon_z/E_f^t$, enters current
 482 parameterizations (i.e., St Laurent et al. 2002) for diapycnal mixing – tightly coupled to internal
 483 wave breaking – in general circulation models. Its value is often assumed constant and set to 0.3,
 484 although there is compelling evidence for strong geographical heterogeneity (q has been reported
 485 to vary from 0.05 to 0.60, see the review in MacKinnon et al. 2017). As the ocean stratification
 486 and the global overturning circulation are highly sensitive to diapycnal mixing (Mashayek et al.
 487 2015; de Lavergne et al. 2016), understanding the physics underpinning the regional variability in
 488 q is important (MacKinnon et al. 2017). Here, a regional estimate of q on the northern MAR is
 489 0.49 ± 0.35 (mean and std dev) for the 9 stations on the ridge (yellow dots in Fig. 12-inset map).
 490 Notice that this estimate applies to the top of the MAR and takes into account E_f^t computed on
 491 a $1/4^\circ$ grid, thus it is tight to a length scale of roughly 25 km. Our regional q is considerably
 492 higher than the 8-25% estimated in Hawaii (Klymak et al. 2006), consistent with an enhanced
 493 generation of high-mode internal tides on the MAR. However, this estimate of q must be inter-
 494 preted cautiously, due to the relatively modest number of dissipation measurements and the their
 495 poorly constrained representativeness – namely linked to the spring-neap variability of dissipation
 496 (section 4c). Additional measurements with a greater spatio-temporal coverage would be needed
 497 to refine this estimate.

6. Conclusions

A multi-source analysis of the lifecycle of semidiurnal internal tides on the MAR sector south of the Azores has been conducted. The main conclusions are:

1. Mooring data on top of the MAR reveal that the internal tide horizontal energy flux is dominated by mode 1, which is steady in intensity and direction (to the south-east). The mode-1 horizontal energy flux undergoes a strong spring-neap cycle that likely stems from interaction with remotely generated internal tides. Energy fluxes for modes greater than 1 are extremely variable in intensity and direction, probably due to interactions with ubiquitous, distributed sources on the MAR.
2. Energy density is more widely distributed among the modes. Specifically, modes 2-10 contain more energy than mode 1 alone. High-mode generation is supported by spectral estimates of energy conversion.
3. Estimates of modal group velocity indicate that most modes are compatible with standing internal waves. Given conclusion 2, this implies that energy is concentrated above the MAR and ultimately dissipates locally. This is supported by the strong energy dissipation inferred from microstructure measurements.
4. A simplified regional energy budget outlines qualitative differences with the well-studied Hawaiian Ridge system (Merrifield and Holloway 2002; Klymak et al. 2006), which dissipates a similar amount of semidiurnal barotropic tide energy (16 GW over the MAR vs. 20 GW around Hawaii). Namely, only 9% (vs. 30% in Hawaii) of the energy is converted into mode 1, the only mode that may radiate energy away. Consistently, the fraction of energy locally dissipated is higher over the MAR, $q = 0.49 \pm 0.35$ vs. $0.08 - 0.25$ in Hawaii (Klymak et al. 2006). This measure is, however, rather uncertain given the modest number of direct

dissipation measurements. Note that these results are in line with differences in internal tide characteristics between the two systems highlighted in St Laurent and Nash (2004). Falahat et al. (2014a) also found a higher q in the Atlantic Ocean than in the Pacific Ocean. They attribute this difference to the extended sharp topography of the MAR as compared to the knife-edge shapes of the Hawaiian Ridge and isolated seamounts in the Pacific Ocean.

A final perspective of this work is provided by the regional validation of the spectral estimate of energy conversion, which can be extended globally. This model is, by construction, more accurate than parameterizations (e.g., Nycander 2005; Green and Nycander 2013, and references therein) and gives additional information on the modal content and direction of the internal tide energy flux.

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References

Alford, M. H., 2003: Redistribution of energy available for ocean mixing by long-range propagation of internal waves. *Nature*, **423** (6936), 159–162, doi:10.1038/nature01628.

543 Alford, M. H., M. C. Gregg, and M. A. Merrifield, 2006: Structure, propagation, and mixing of
 544 energetic baroclinic tides in Mamala Bay, Oahu, Hawaii. *J. Phys. Oceanogr.*, **36** (6), 997–1018,
 545 doi:10.1175/JPO2877.1.

546 Alford, M. H., and Z. Zhao, 2007a: Global patterns of low-mode internal-wave propagation. Part
 547 I: Energy and energy flux. *J. Phys. Oceanogr.*, **37** (7), 1829–1848, doi:10.1175/JPO3085.1.

548 Alford, M. H., and Z. Zhao, 2007b: Global patterns of low-mode internal-wave propagation. Part
 549 II: Group velocity. *J. Phys. Oceanogr.*, **37** (7), 1849–1858, doi:10.1175/JPO3086.1.

550 Ansong, J. K., and Coauthors, 2017: Semidiurnal internal tide energy fluxes and their variability in
 551 a Global Ocean Model and moored observations. *J. Geophys. Res. Oceans*, **122** (3), 1882–1900,
 552 doi:10.1002/2016JC012184.

553 Armi, L., 1979: Effects of variations in eddy diffusivity on property distributions in the oceans. *J.*
 554 *Mar. Res.*, **37** (3).

555 Baines, P. G., 1998: *Topographic effects in stratified flows*. Cambridge Univ. Press, 482 pp.

556 Becker, J., and Coauthors, 2009: Global bathymetry and elevation data at 30 arc seconds resolu-
 557 tion: SRTM30_PLUS. *Marine Geodesy*, **32** (4), 355–371, doi:10.1080/01490410903297766.

558 Bell, T., 1975a: Lee waves in stratified flows with simple harmonic time dependence. *J. Fluid*
 559 *Mech.*, **67** (04), 705–722, doi:10.1017/S0022112075000560.

560 Bell, T., 1975b: Topographically generated internal waves in the open ocean. *J. Geophys. Res.*,
 561 **80** (3), 320–327.

562 Dalcin, L., R. Paz, M. Storti, and J. D’Elia, 2008: MPI for Python: Performance improvements
 563 and MPI-2 extensions. *Journal of Parallel and Distributed Computing*, **68** (5), 655–662, doi:
 564 10.1016/j.jpdc.2007.09.005.

565 Dalcin, L. D., R. R. Paz, P. A. Kler, and A. Cosimo, 2011: Parallel distributed computing using
 566 Python. *Advances in Water Resources*, **34** (9), 1124–1139, doi:10.1016/j.advwatres.2011.04.
 567 013.

568 de Lavergne, C., G. Madec, J. Le Sommer, A. G. Nurser, and A. C. Naveira Garabato, 2016: The
 569 impact of a variable mixing efficiency on the abyssal overturning. *J. Phys. Oceanogr.*, **46** (2),
 570 663–681, doi:10.1175/JPO-D-14-0259.1.

571 Desaubies, Y., and M. Gregg, 1981: Reversible and irreversible finestructure. *J. Phys. Oceanogr.*,
 572 **11** (4), 541–556.

573 Dunphy, M., A. L. Ponte, P. Klein, and S. Le Gentil, 2017: Low-mode internal tide propagation in
 574 a turbulent eddy field. *J. Phys. Oceanogr.*, **47** (3), 649–665, doi:10.1175/JPO-D-16-0099.1.

575 Egbert, G., and R. Ray, 2000: Significant dissipation of tidal energy in the deep ocean inferred
 576 from satellite altimeter data. *Nature*, **405** (6788), 775–778, doi:10.1038/35015531.

577 Egbert, G. D., and S. Y. Erofeeva, 2002: Efficient inverse modeling of barotropic ocean tides. *J.*
 578 *Atmos. Ocean. Technol.*, **19** (2), 183–204, doi:10.1175/1520-0426(2002)019<0183:EIMOBO>
 579 2.0.CO;2.

580 Egbert, G. D., and R. D. Ray, 2001: Estimates of M2 tidal energy dissipation from
 581 TOPEX/Poseidon altimeter data. *J. Geophys. Res.*, **106** (C10), 22 475–22 502, doi:10.1029/
 582 2000JC000699.

583 Falahat, S., J. Nycander, F. Roquet, A. M. Thurnherr, and T. Hibiya, 2014a: Comparison of calcu-
 584 lated energy flux of internal tides with microstructure measurements. *Tellus A*, **66** (1), 23 240,
 585 doi:10.3402/tellusa.v66.23240.

586 Falahat, S., J. Nycander, F. Roquet, and M. Zarroug, 2014b: Global calculation of tidal energy
 587 conversion into vertical normal modes. *J. Phys. Oceanogr.*, **44** (12), 3225–3244, doi:10.1175/
 588 JPO-D-14-0002.1.

589 Garrett, C., and E. Kunze, 2007: Internal tide generation in the deep ocean. *Annu. Rev. Fluid*
 590 *Mech.*, **39**, 57–87, doi:10.1146/annurev.fluid.39.050905.110227.

591 Gill, A. E., 1982: *Atmosphere-ocean dynamics*. Academic Press, 662 pp.

592 Goff, J. A., 1991: A global and regional stochastic analysis of near-ridge abyssal hill morphology.
 593 *J. Geophys. Res.*, **96** (B13), 21 713–21 737.

594 Green, J. M., and J. Nycander, 2013: A comparison of tidal conversion parameterizations for tidal
 595 models. *J. Phys. Oceanogr.*, **43** (1), 104–119, doi:10.1175/JPO-D-12-023.1.

596 Jayne, S. R., and L. C. St Laurent, 2001: Parameterizing tidal dissipation over rough topography.
 597 *Geophys. Res. Lett.*, **28** (5), 811–814, doi:10.1029/2000GL012044.

598 Kerry, C. G., B. S. Powell, and G. S. Carter, 2014: The impact of subtidal circulation on internal-
 599 tide-induced mixing in the Philippine Sea. *J. Phys. Oceanogr.*, **44** (12), 3209–3224, doi:10.
 600 1175/JPO-D-13-0249.1.

601 Klymak, J. M., R. Pinkel, and L. Rainville, 2008: Direct breaking of the internal tide near topogra-
 602 phy: Kaena Ridge, Hawaii. *J. Phys. Oceanogr.*, **38** (2), 380–399, doi:10.1175/2007JPO3728.1.

603 Klymak, J. M., and Coauthors, 2006: An estimate of tidal energy lost to turbulence at the Hawaiian
 604 Ridge. *J. Phys. Oceanogr.*, **36** (6), 1148–1164, doi:10.1175/JPO2885.1.

605 Kunze, E., 2017: Internal-wave-driven mixing: Global geography and budgets. *J. Phys. Oceanogr.*,
 606 **in press**, doi:10.1175/JPO-D-16-0141.1.

607 Ledwell, J., E. Montgomery, K. Polzin, L. St Laurent, R. Schmitt, and J. Toole, 2000: Evidence
 608 for enhanced mixing over rough topography in the abyssal ocean. *Nature*, **403** (6766), 179–182,
 609 doi:10.1038/35003164.

610 Lefauve, A., C. Muller, and A. Melet, 2015: A three-dimensional map of tidal dissipation over
 611 abyssal hills. *J. Geophys. Res. Oceans*, **120** (7), 4760–4777, doi:10.1002/2014JC010598.

612 Llewellyn Smith, S. G., and W. Young, 2002: Conversion of the barotropic tide. *J. Phys.*
 613 *Oceanogr.*, **32** (5), 1554–1566, doi:10.1175/1520-0485(2002)032<1554:COTBT>2.0.CO;2.

614 Locarnini, R., and Coauthors, 2013: World Ocean Atlas 2013, Volume 1: Temperature. *NOAA*
 615 *Atlas NESDIS*, **73**, 40, URL <http://www.nodc.noaa.gov/OC5/indprod.htm>.

616 Lueck, R. G., and T. D. Mudge, 1997: Topographically induced mixing around a shallow
 617 seamount. *Science*, **276** (5320), 1831–1833.

618 MacKinnon, J. A., and Coauthors, 2017: Climate process team on internal-wave driven ocean
 619 mixing. *Bull. Amer. Meteor. Soc.*, **in press**, doi:10.1175/BAMS-D-16-0030.1.

620 Mashayek, A., R. Ferrari, M. Nikurashin, and W. Peltier, 2015: Influence of enhanced abyssal
 621 diapycnal mixing on stratification and the ocean overturning circulation. *J. Phys. Oceanogr.*,
 622 **45** (10), 2580–2597, doi:10.1175/JPO-D-15-0039.1.

623 Melet, A., M. Nikurashin, C. Muller, S. Falahat, J. Nycander, P. G. Timko, B. K. Arbic, and J. A.
 624 Goff, 2013: Internal tide generation by abyssal hills using analytical theory. *J. Geophys. Res.*
 625 *Oceans*, **118** (11), 6303–6318, doi:10.1002/2013JC009212.

626 Merrifield, M. A., and P. E. Holloway, 2002: Model estimates of M2 internal tide energetics at the
 627 Hawaiian Ridge. *J. Geophys. Res.*, **107** (C8), doi:10.1029/2001JC000996.

628 Munk, W., and C. Wunsch, 1998: Abyssal recipes II: energetics of tidal and wind mixing. *Deep*
629 *Sea Res. Part I*, **45 (12)**, 1977–2010.

630 Nash, J., M. Alford, E. Kunze, K. Martini, and S. Kelly, 2007: Hotspots of deep ocean mixing on
631 the Oregon continental slope. *Geophys. Res. Lett.*, **34 (1)**, doi:10.1029/2006GL028170.

632 Nash, J. D., M. H. Alford, and E. Kunze, 2005: Estimating internal wave energy fluxes in the
633 ocean. *J. Atmos. Oceanic Technol.*, **22 (10)**, 1551–1570, doi:10.1175/JTECH1784.1.

634 Nash, J. D., E. Kunze, J. M. Toole, and R. W. Schmitt, 2004: Internal tide reflection and tur-
635 bulent mixing on the continental slope. *J. Phys. Oceanogr.*, **34 (5)**, 1117–1134, doi:10.1175/
636 1520-0485(2004)034<1117:ITRATM>2.0.CO;2.

637 Nycander, J., 2005: Generation of internal waves in the deep ocean by tides. *J. Geophys. Res.*,
638 **110 (C10)**, doi:10.1029/2004JC002487.

639 Oakey, N., 1982: Determination of the rate of dissipation of turbulent energy from simultaneous
640 temperature and velocity shear microstructure measurements. *J. Phys. Oceanogr.*, **12 (3)**, 256–
641 271.

642 Polzin, K., J. Toole, J. Ledwell, and R. Schmitt, 1997: Spatial variability of turbulent mixing in
643 the abyssal ocean. *Science*, **276 (5309)**, 93–96, doi:10.1126/science.276.5309.93.

644 Rainville, L., and R. Pinkel, 2006: Propagation of low-mode internal waves through the ocean. *J.*
645 *Phys. Oceanogr.*, **36 (6)**, 1220–1236, doi:10.1175/JPO2889.1.

646 Ray, R., 1998: Ocean self-attraction and loading in numerical tidal models. *Marine Geodesy*,
647 **21 (3)**, 181–192.

648 Rudnick, D. L., and Coauthors, 2003: From tides to mixing along the Hawaiian Ridge. *Science*,
649 **301 (5631)**, 355–357, doi:10.1126/science.1085837.

650 Searle, R., 1987: Regional setting and geophysical characterization of the Great Meteor East area
 651 in the Madeira Abyssal Plain. *Geological Society, London, Special Publications*, **31 (1)**, 49–70.

652 Sharples, J., 2016: Cruise Report RRS James Clark Ross JR15-007 (RidgeMix). Tech. rep., Uni-
 653 versity of Liverpool, UK, 104 pp.

654 Smith, W., and D. Sandwell, 1997: Global sea floor topography from satellite altimetry and ship
 655 depth soundings. *Science*, **277 (5334)**, 1956–1962, doi:10.1126/science.277.5334.1956.

656 St. Laurent, L., and C. Garrett, 2002: The role of internal tides in mixing the deep ocean. *J. Phys.*
 657 *Oceanogr.*, **32 (10)**, 2882–2899, doi:10.1175/1520-0485(2002)032<2882:TROITI>2.0.CO;2.

658 St Laurent, L., and J. Nash, 2004: An examination of the radiative and dissipative properties of
 659 deep ocean internal tides. *Deep Sea Res. II*, **51 (25)**, 3029–3042, doi:10.1016/j.dsr2.2004.09.
 660 008.

661 St Laurent, L., H. Simmons, and S. Jayne, 2002: Estimating tidally driven mixing in the deep
 662 ocean. *Geophys. Res. Lett.*, **29 (23)**, doi:10.1029/2002GL015633.

663 Timko, P. G., B. K. Arbic, J. A. Goff, J. K. Ansong, W. H. Smith, A. Melet, and A. J. Wallcraft,
 664 2017: Impact of synthetic abyssal hill roughness on resolved motions in numerical global ocean
 665 tide models. *Ocean Modell.*, **112**, 1–16, doi:10.1016/j.ocemod.2017.02.005.

666 Wunsch, C., and R. Ferrari, 2004: Vertical mixing, energy, and the general circulation of the
 667 oceans. *Annu. Rev. Fluid Mech.*, **36**, 281–314, doi:10.1146/annurev.fluid.36.050802.122121.

668 Zaron, E. D., and G. D. Egbert, 2014: Time-variable refraction of the internal tide at the Hawaiian
 669 Ridge. *J. Phys. Oceanogr.*, **44 (2)**, 538–557, doi:10.1175/JPO-D-12-0238.1.

670 Zhao, Z., M. H. Alford, J. B. Girton, L. Rainville, and H. L. Simmons, 2016: Global observations
671 of open-ocean mode-1 M_2 internal tides. *J. Phys. Oceanogr.*, **46** (6), 1657–1684, doi:10.1175/
672 JPO-D-15-0105.1.

673 Zhao, Z., M. H. Alford, J. A. MacKinnon, and R. Pinkel, 2010: Long-range propagation of the
674 semidiurnal internal tide from the Hawaiian Ridge. *J. Phys. Oceanogr.*, **40** (4), 713–736, doi:
675 10.1175/2009JPO4207.1.

676 Zilberman, N., J. Becker, M. Merrifield, and G. Carter, 2009: Model estimates of M_2 internal
677 tide generation over Mid-Atlantic Ridge topography. *J. Phys. Oceanogr.*, **39** (10), 2635–2651,
678 doi:10.1175/2008JPO4136.1.

679 Zweng, M., and Coauthors, 2013: World Ocean Atlas 2013, Volume 2: Salinity. *NOAA Atlas*
680 *NESDIS*, **74**, 39, URL <http://www.nodc.noaa.gov/OC5/indprod.htm>.

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mode	1	2	3	4	5	6	7	8	9	10
E (kJ m ⁻²)	0.84±0.25	0.26±0.16	0.23±0.13	0.16±0.12	0.11±0.08	0.07±0.04	0.05±0.03	0.05±0.04	0.04±0.02	0.05±0.03
F (kW m ⁻¹)	0.83±0.25	0.07±0.05	0.05±0.04	0.03±0.02	0.01±0.01	0.01±0.01	0.01±0.00	–	–	–
c_g^m (m s ⁻¹)	1.09±0.10	0.34±0.05	0.33±0.04	0.33±0.03	0.23±0.03	0.19±0.03	0.16±0.01	0.11±0.01	0.06±0.02	0.05±0.01
c_g (m s ⁻¹)	1.33	0.66	0.44	0.33	0.27	0.22	0.19	0.17	0.15	0.13
c_g^s (m s ⁻¹)	1.08	0.54	0.36	0.27	0.22	0.18	0.15	0.14	0.12	0.11

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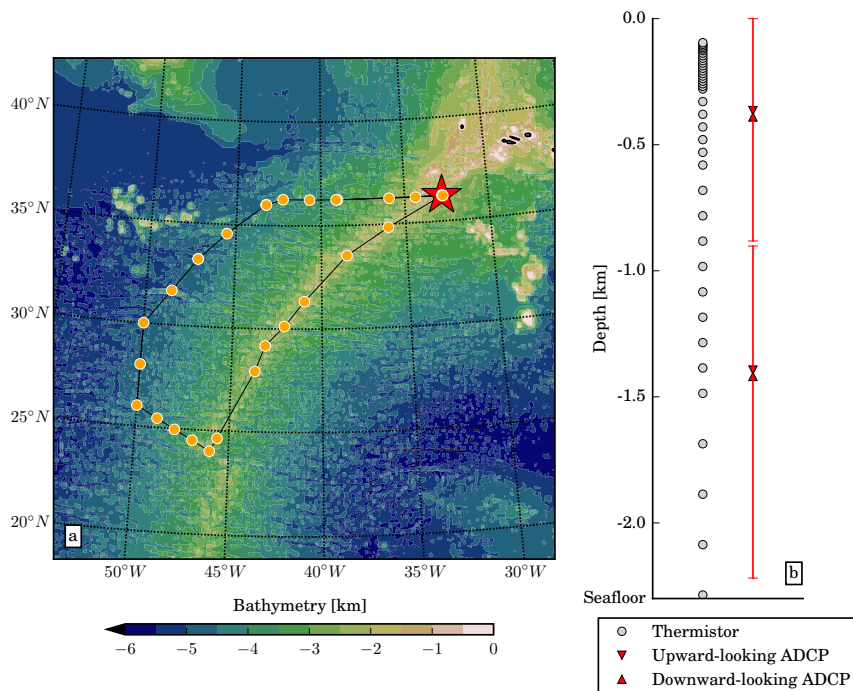


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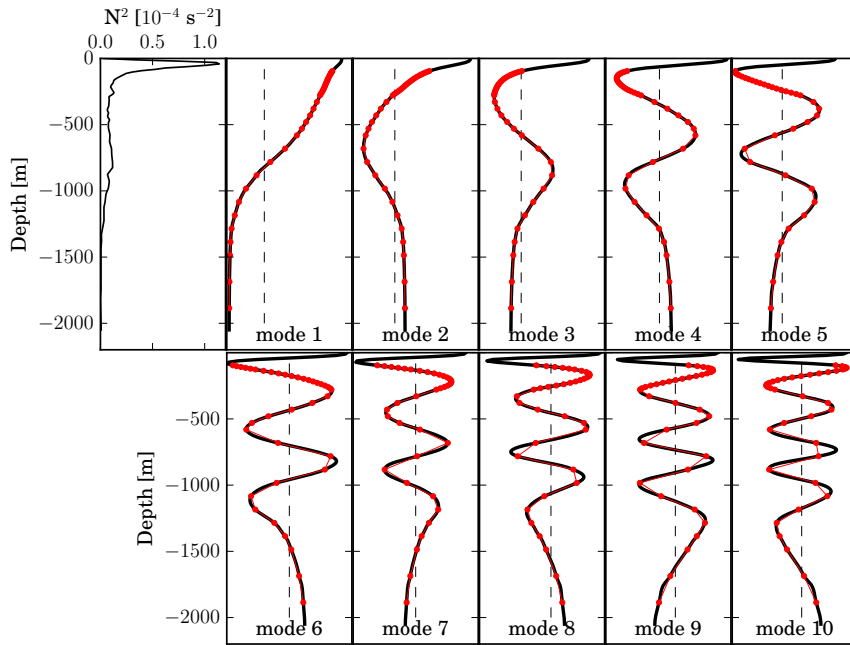


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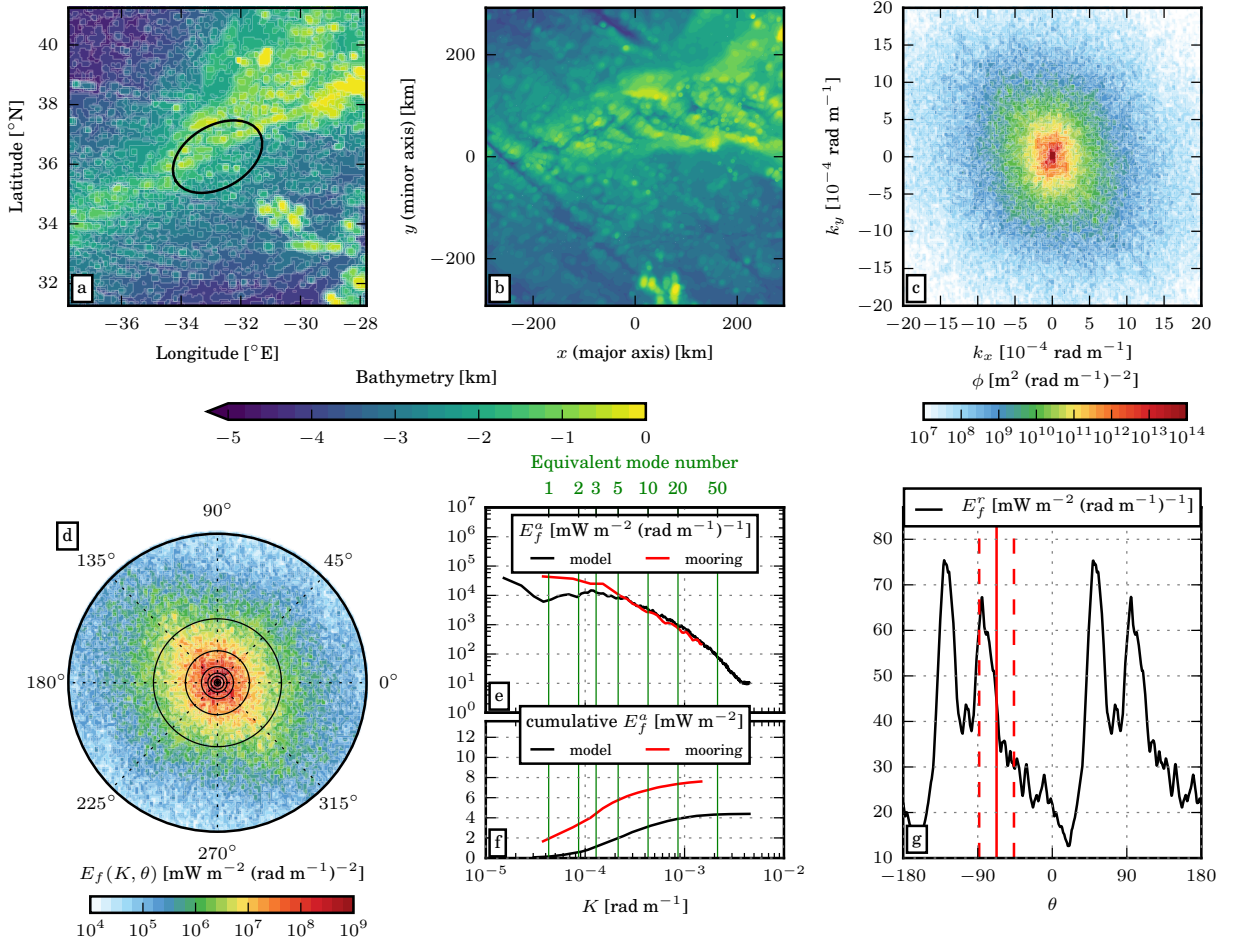
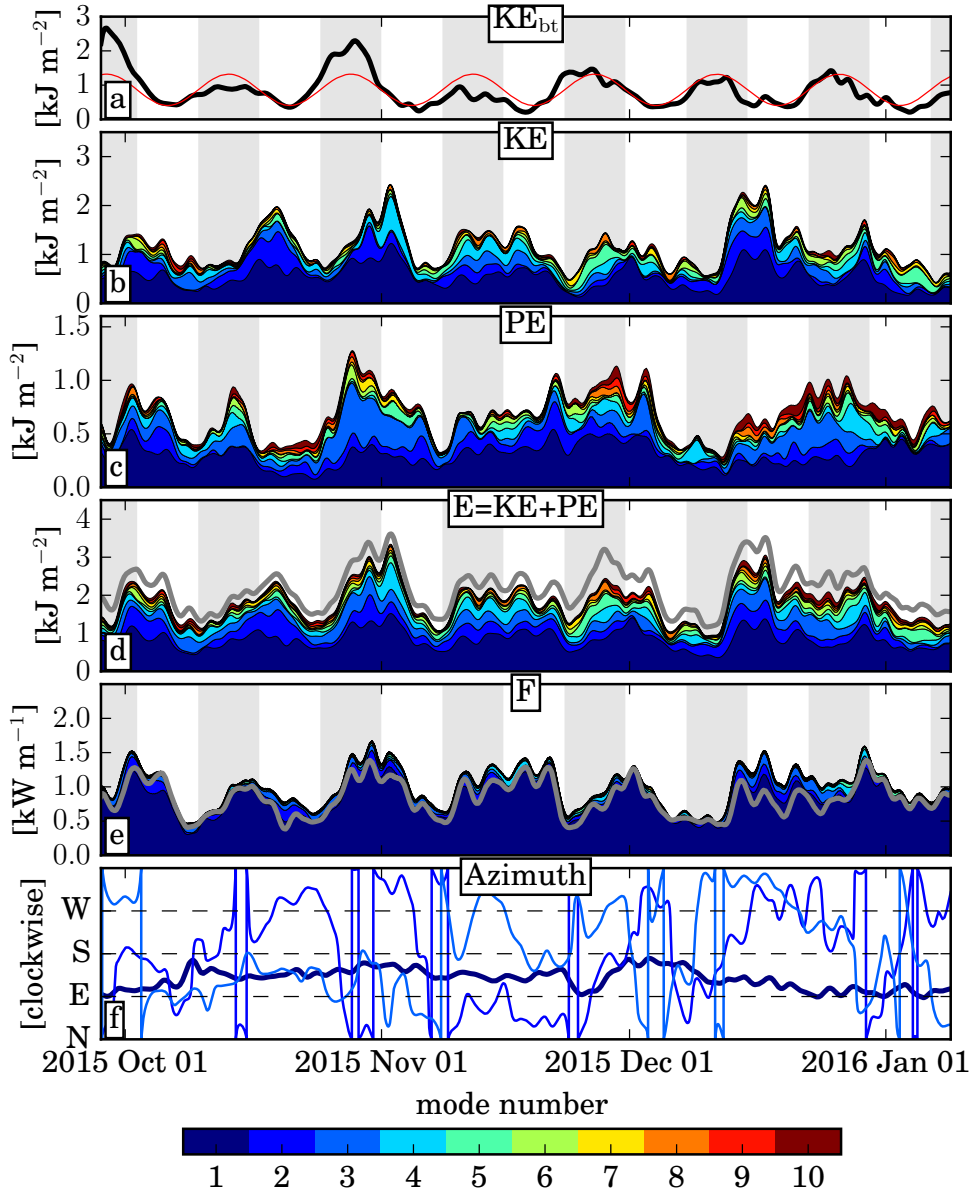


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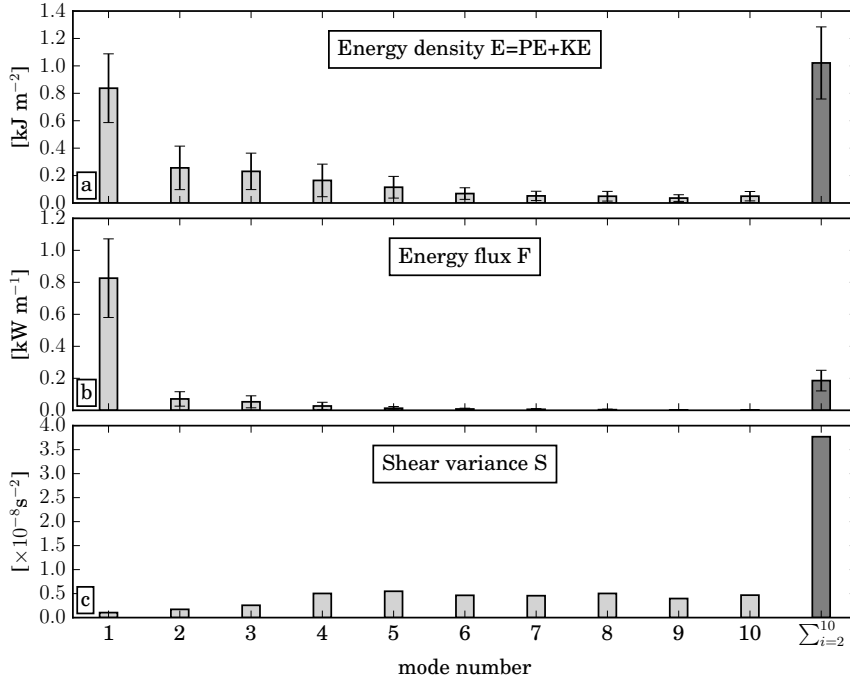


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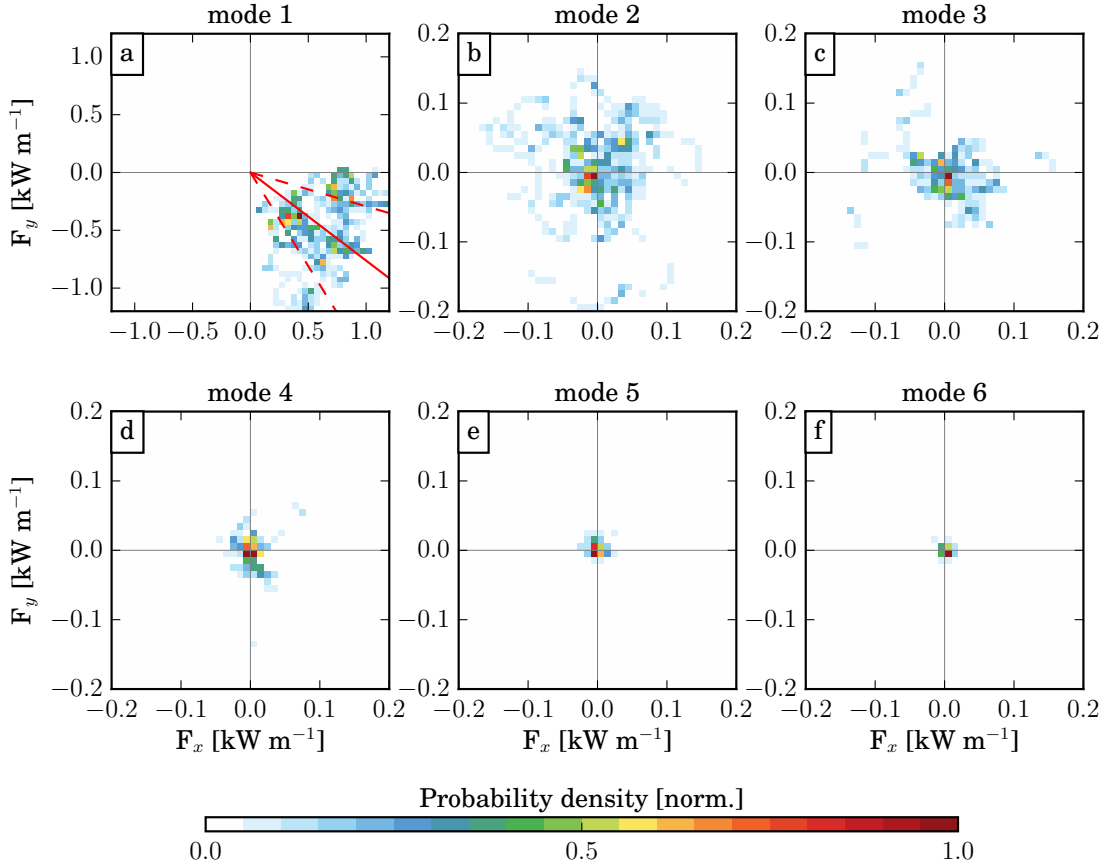


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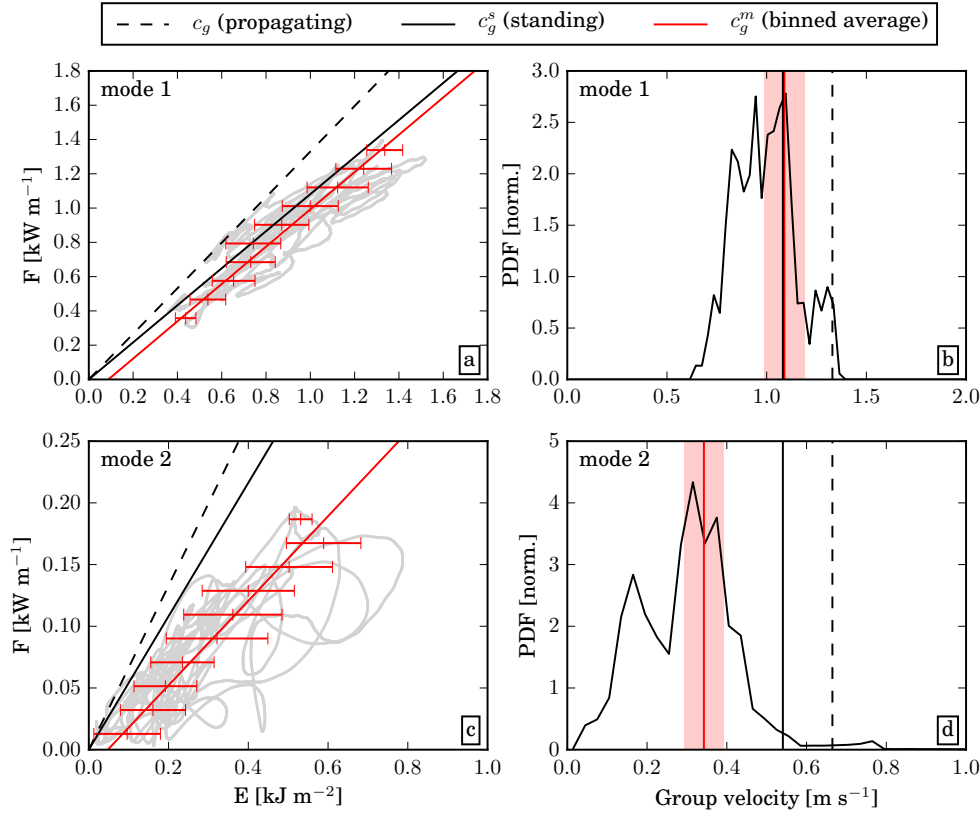
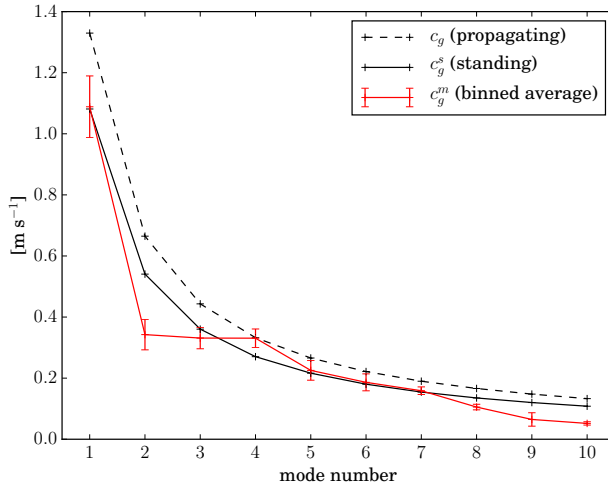


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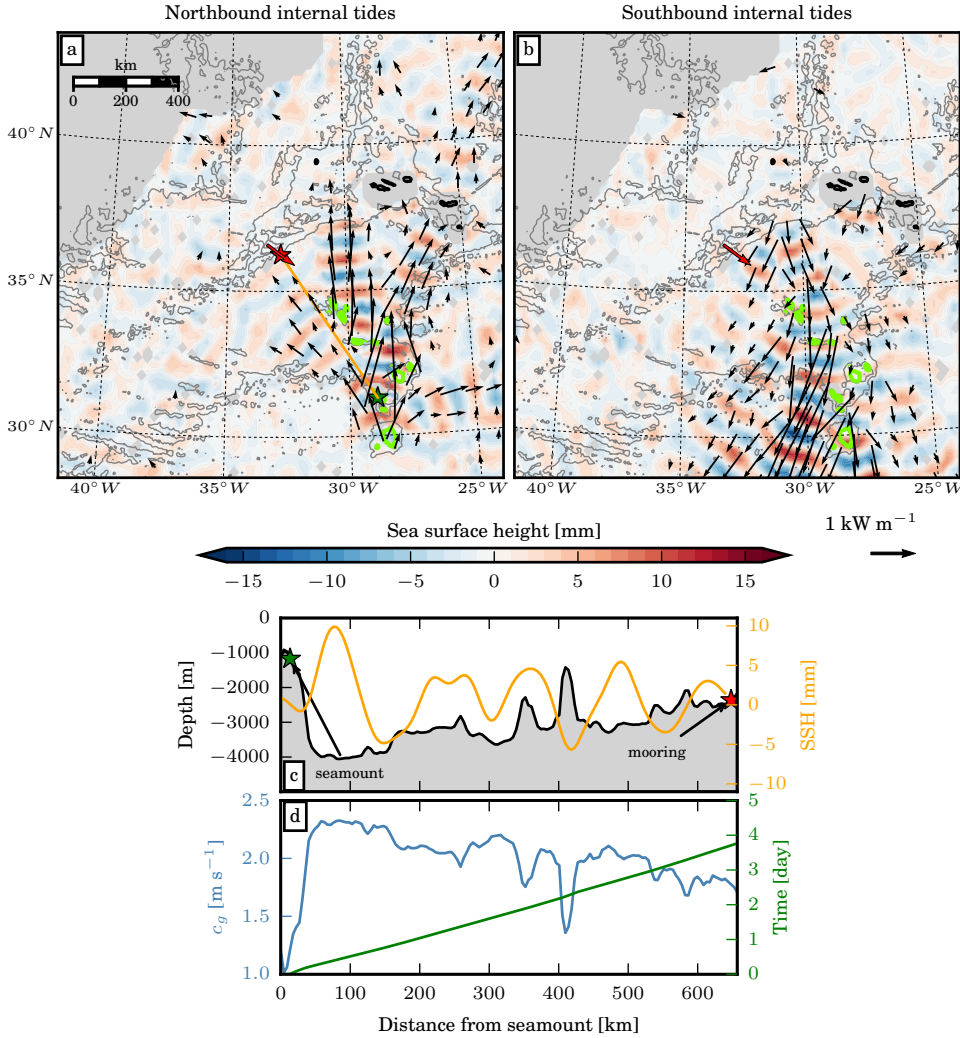
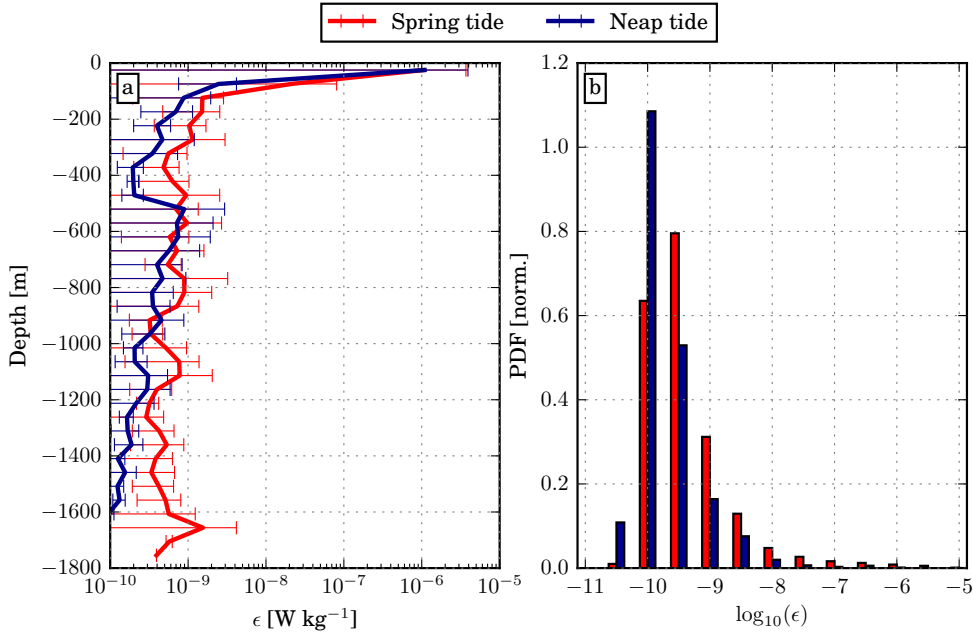


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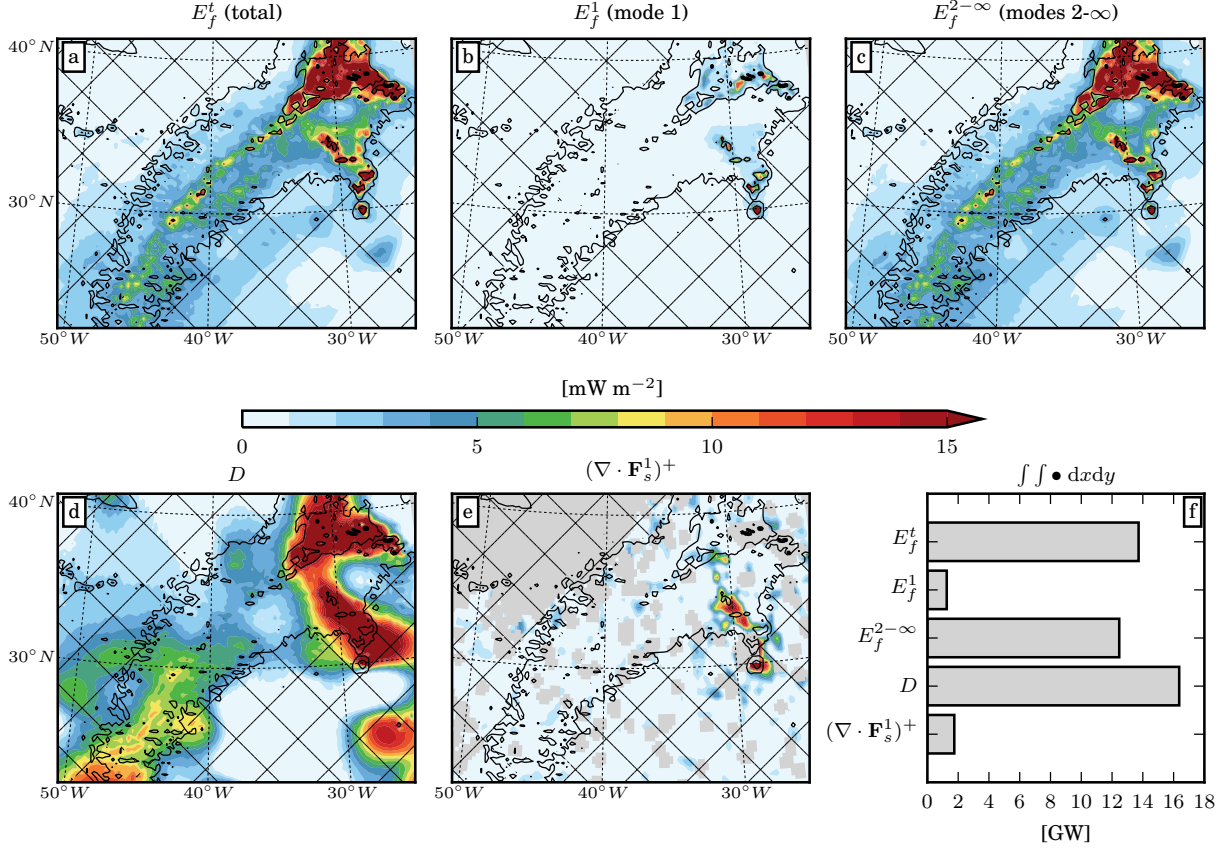


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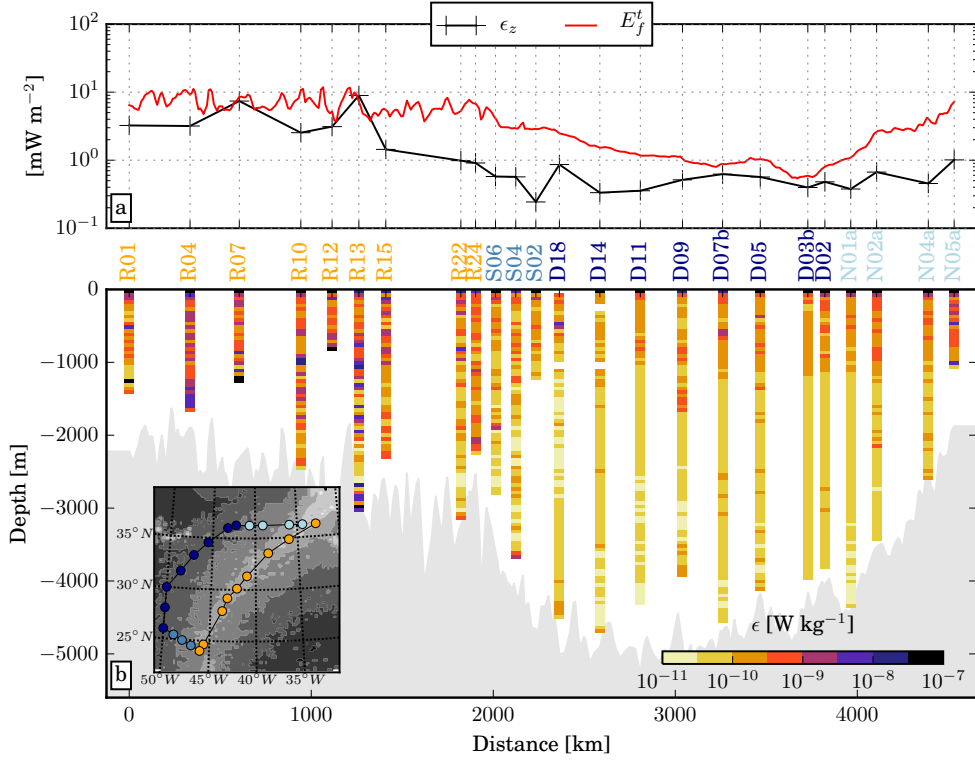


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